Philip E. van Beynen, Henry P. Schwarcz, and Derek C. Ford - Holocene climatic variation recorded in a speleothem from McFail's Cave, New York. *Journal of Cave and Karst Studies,* v. 66, no. 1, p. 20-27.

HOLOCENE CLIMATIC VARIATION RECORDED IN A SPELEOTHEM FROM MCFAIL'S CAVE, NEW YORK

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A speleothem collected from McFail's Cave, central New York, was analyzed in order to produce a high resolution paleoclimatic record for this region. The record is dated by Th/U ICP Mass Spectrometry. Variation in growth rates and $\delta^{18}O$ values for the period 0 to 7.6 ka revealed three distinct intervals: maximum warmth and wettest from 7.6 to 7.0 ka; a slow steady cooling from 7.0 to 2.5 ka; and fairly constant temperatures for the last few thousand years. The climatic optimum appears to have occurred at or before 7.6 ka. These changes are in agreement with regional stable isotopic and lake level records from the Finger Lakes in NY, suggesting the climatic changes are regional in scope. Enrichment of ^{18}O in speleothems deposited at this time suggests that temperatures could potentially have been 5°C warmer than at present, which is consistent with pollen records for this region. However, a northward shift in the Jet Stream during the period would have affected the source area of the rainfall, providing heavier $\delta^{18}O$ values, an effect that would potentially reduce our temperature increase of 5°C. Quantification of this shift's contribution to the temperature increase recorded by our speleothem is extremely difficult due to the lack of information about the exact location of the Jet Stream. $\delta^{13}C$ values are generally uniform but show a brief vegetational optimum at ~7.5 ka and increased density from 5.5 to 2.5 ka, suggesting a wetter climate.

Calcite speleothems precipitated in caves have been shown to record paleo-environmental change through the Late Quaternary and Holocene periods (Harmon *et al*. 1978; Gascoyne 1992; Bar-Matthews *et al*. 1999). Stable isotopic analysis of microgram samples of calcite and their fluid inclusions, coupled with accurate U/Th dating techniques, may provide high-resolution records of the climate and vegetation changes above a cave (Schwarcz 1986). Paleoclimatic inferences may be drawn by measuring growth rates of millennial (Hennig *et al*. 1983) to annual scale (Broecker *et al*. 1960; Baker *et al*. 1993; Shopov *et al*. 1994; 1997; Genty & Quinif 1996; Roberts *et al*. 1998; van Beynen 1998; Lauritzen *et al*. 1999; Perrette *et al*. 1999).

As water from rain and snow melt percolates through the soil, it absorbs soil CO_2 produced by root respiration, increasing its acidity. On reaching limestone bedrock beneath, the acidic water dissolves calcium carbonate until thermodynamic equilibrium with the ambient partial pressure of CO_2 (P_{CO2}) is attained (Plummer *et al*. 1979). On entering an airfilled cave, the CO_2 may degas rapidly because P_{CO2} is usually lower than in the soil and tiny fissures above it; precipitation of part of the dissolved $CaCO_3$ then results (Plummer *et al*. 1979). Where calcite has formed in oxygen isotopic equilibrium with ambient water, the isotopic fractionation between calcite and water, $\Delta_{c\text{-}w}$, is dependent on the temperature (O'Neill *et al*. 1975):

$$1000 \ln \Delta_{c\text{-}w} = 2.78 \times 10^6\, T^{-2}\, (K^{-2}) - 2.89 \qquad [1]$$

Where $\Delta_{c\text{-}w} = (1000 + \delta^{18}O_c) / (1000 + \delta^{18}O_w)$ and $\delta^{18}O_c$ and $\delta^{18}O_w$ are the $\delta^{18}O$ values of calcite and water, respectively and K is the equilibrium constant. We can test for equilibrium deposition by analysis of multiple samples taken from a single growth layer. Equilibrium deposits display the following characteristics: a) there is no variation in $\delta^{18}O$ along the growth layer; b) there is no correlation between $\delta^{18}O$ and $\delta^{13}C$ (Hendy & Wilson 1968). Deep interior cave temperatures are usually invariant over the year and close to the mean annual surface temperature above the cave (Wigley & Brown 1976). The oxygen isotope records of speleothems may therefore offer proxy records of paleotemperatures and climate change (Schwarcz 1986). In practice, however, it is difficult to use analysis of speleothems to determine the temperature of deposition (T) because the relation between T and $\delta^{18}O_c$ is principally controlled by two factors which have opposing T-dependencies: the isotopic fractionation between calcite and water decreases with increasing T by about 0.24 ‰/°C; but $\delta^{18}O$ (ppt), the isotopic composition of the meteoric precipitation, increases with increasing T by between 0.3 and 0.7 ‰/°C, a rate which varies locally and with time as a consequence of differing sources and storm tracks (Schwarcz 1986). The balance between these factors (as well as others, discussed in Schwarcz 1986) may be such that the T-dependence of $\delta^{18}O_c$ ($\gamma = d\, \delta^{18}O_c/dT$) may vary in magnitude and sign from one site to another. In general we expect that at a given locality the sign of γ will remain the same through time (Frumkin *et al*. 1999).

It is possible in theory to determine T directly by simultaneous analysis of the calcite and of water trapped as fluid inclusions in it (Schwarcz *et al*. 1976). It is also useful to use $\delta^{18}O_c$ by itself as a proxy record of climate. To do so we must,

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however, determine the sign of γ within the speleothem in question. We may do this by comparison of $\delta^{18}O$ of modern calcite with that of Ice Age speleothems that are inferred to have formed at lower temperatures. Similarly, for post-Ice Age deposits (the Holocene–the last 11,000 years), we can assume that deposits dating to a well established mid-Holocene warm phase (the Hypsithermal) were deposited at warmer temperatures than modern calcite. The gradient in $\delta^{18}O$ between calcite formed in these differing periods defines the sign of γ.

The $\delta^{13}C$ record in speleothems is a function of the vegetation density and types of plants growing above the cave. The proportion of dissolved inorganic carbon in recharge water derived from plant respiration increases with plant density and plant biological activity. $^{13}C/^{12}C$ ratios in speleothems represent a mixture of carbon-derived soil CO_2 and of the limestone bedrock (~0‰). Drip rates, length of flow path, and rates of outgassing and calcite precipitation may also affect the $\delta^{13}C$ signal (Wigley *et al*. 1978). Where $\delta^{13}C$ of plants has remained constant, lower $\delta^{13}C$ (lighter) values indicate more forested environments and/or greater respiratory activity in warmer, wetter conditions compared to the higher $\delta^{13}C$ (heavier) values beneath more open, less humid vegetative environments. Frumkin *et al*. (1999), in a study of speleothems from Nahal Qanah Cave, Israel, showed that more depleted $\delta^{13}C$ values in a Holocene speleothem correlated with increased density of vegetation there.

Most previous research on the paleoclimate of the Holocene epoch in the northeastern USA has produced fairly broad interpretations often with only brief snapshots in time. The dates used in these below papers were originally radiocarbon dates, but as our Th/U dating is in calendar years, we have converted the original dates of the articles into calendar years using the INTCAL98 conversion data from Stuiver *et al*. (1998). This allows direct comparison between our data and that from the below studies. Prentice *et al*. (1991) gave 3000-year increments for the Holocene using pollen data. They concluded that the climate at 6.9 to 10.2 ka (ka = thousand years) was warmer and wetter than present, and that after 3.5 ka BP (BP – before present) summer temperatures declined. From studies of past distributions of tree species, Davis *et al*. (1980) concluded that this region was warmer before 5.7 ka. Jackson and Whitehead (1991) highlighted three periods from their pollen records; a) 10.2 to 5.7 ka: summer temperatures 1.0 to 1.6°C higher than present; b) 5.7 to 3.5 ka: transition to more modern forest assemblages; and c) after 3.5 ka: modern vegetation present. Lacustrine sediments from the Finger Lakes Region of New York show high lake levels at ~6.3 to 9.5 ka with maximum levels at 7.8 ka (Dywer *et al*. 1996). Studies of Cayuga Lake, NY, sediments show a warm period centered on ~8.9 ka (Anderson *et al*. 1997); and high $CaCO_3$ accumulation rates (Mullins 1998). Therefore, there is some debate as to the precise timing of the Hypsithermal in this region, with estimates ranging between 7.8 ka and 8.9 ka BP.

In this paper we have used stable oxygen and carbon isotopic records from a speleothem collected from McFail's Cave, New York to establish the most detailed isotopic record for the northeastern USA to date. We will also determine how our Holocene climate record for the NY region compares to other proxy records for the area.

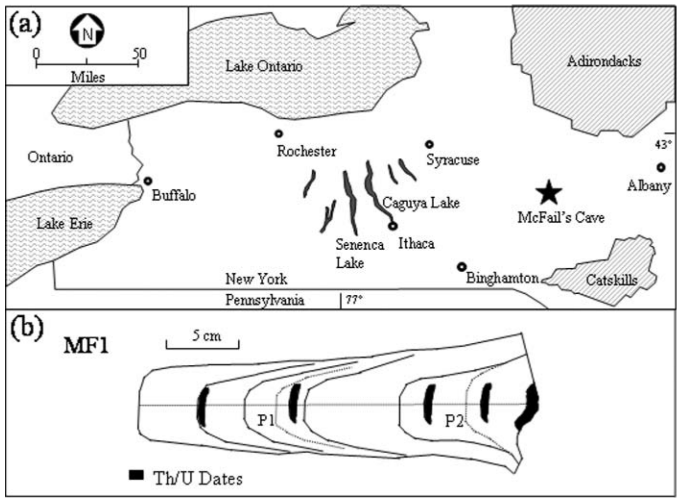

Figure 1. a) Location of McFail's Cave; b) sampling sites for dating; straight dotted line running up the middle of the sample shows the transect taken for stable isotope analysis. P1 and P2 show positions of stable isotope profiles taken for testing for equilibrium (see text).

STUDY AREA

McFail's Cave is located 3 km northeast of Cobleskill (38°22'N, 90°39'W) in Schoharie County, New York (Fig. 1). The sample was obtained from an active river passage that is traceable for 4.5 km at an elevation of 400 to 430 m asl. It is developed in limestone and dolostone beds of the Helderberg Group (Upper Silurian-Lower Devonian) ranging up to 50 m in thickness, and covered by up to 100 m of younger limestones (Palmer, 1975). The average thickness of bedrock above the cave is between 50-70 m.

The region was glaciated during the late Wisconsin, when 50 m of glacial till was deposited in the Cobleskill Valley (Palmer 1975). However, there is only a thin veneer of till above the cave. The vegetation cover includes a small marsh and some fields surrounded by a closed canopy Carolinian forest. The marsh is fed by groundwater from surrounding hillsides and is the source for much of the cave dripwater. The regional climate is temperate with mean monthly temperatures ranging from –4.7°C in January to 19.1°C in July (climatic data from National Climatic Data Center, 2003). The average cave temperature for the year is 7.3°C (A. Palmer pers. comm.). The mean total precipitation is 960 mm and is rather uniformly distributed over the year (NCDC, 2003).

The vegetation above McFail's Cave is dominated by C_3 plants (forest environment as opposed to C_4 plants which are predominately grasses) that normally display $\delta^{13}C$ values averaging around –26‰ (Deines 1980). Pollen studies from this

Table 1. U-series dates for MF1.

mm from Base	U (ppm)	$^{234}U/^{238}U$	$^{230}Th/^{234}U$	$^{230}Th/^{232}Th$	Uncorrected Age (ka)	Corrected Age (ka)
2	1.63	2.194 ± 0.0021	0.069 ± 0.0004	51.26 ± 0.37	7.826 ± 0.047	7.606 ± 0.047
37	1.61	2.244 ± 0.0021	0.068 ± 0.0004	306.76 ± 2.60	7.606 ± 0.057	7.570 ± 0.057
75	1.24	2.250 ± 0.0019	0.066 ± 0.0006	105.58 ± 1.15	7.425 ± 0.077	7.324 ± 0.077
165	0.77	2.256 ± 0.0029	0.046 ± 0.0006	23.35 ± 0.33	5.126 ± 0.072	4.805 ± 0.072
225	1.63	2.317 ± 0.0020	0.022 ± 0.0019	15.42 ± 1.31	2.484 ± 0.214	2.245 ± 0.214

region indicate that past vegetation at this site was also dominantly C_3 (Jackson & Whitehead 1991).

METHODS

SAMPLE COLLECTION

Speleothem MF1 is a stalagmite collected from the downstream end of McFail's Cave. It was previously found to be of Holocene age through U-series dating using alpha spectrometry (Gascoyne 1979). The stalagmite was cut along its growth axis and polished to provide a cross section. Micro-samples of 0.1 mg were drilled (high speed dental drill) along growth layers to test for equilibrium deposition and up the growth axis, providing 136 samples for oxygen and carbon analysis. Calcite samples of 1.0–1.25 g each were taken from the speleothem for Th/U dating as illustrated in Figure 1. Calcite samples for Th/U dating were taken within defined growth layers to maximize the resolution of the dates.

U-SERIES DATING

U-series dating was undertaken using inductively coupled plasma (ICP) mass spectrometry. U and Th were extracted from the calcite samples in a clean lab in McMaster University using the methods of Li *et al*. (1989). The $^{229}Th/^{236}U$ spike was calibrated by J. Lundberg, Carleton University. The Th and U extracts were analyzed by A. Simonetti on a VG IsoProbe multicollector ICP mass spectrometer at the Center for Research in Geochemistry and Geodynamics (GEOTOP), University of Quebec at Montreal, using the following protocol.

Samples were analyzed in static multicollection mode using four Faraday collectors (high mass side). Typical settings for the ICP source are the following: Forward power- 1350 watts; Cool gas- 13.50 L/min. (Ar); Intermediate gas- 1.030 L/min. (Ar); Nebulising gas- 0.8–0.9 L/min. (Ar). The instrument settings were optimized using an ICP-MS-grade uranium solution. Samples were conditioned in 2%HNO_3 and introduced into the ICP source using an Aridus© microconcentric nebuliser at an uptake rate of ~50µl/min (verified periodically). The sweep gas settings for the Aridus© introduction system typically varied between 3.5–4.0 L/min argon and 10–15 ml/min nitrogen. Oxide levels are systematically verified and are typically <1% of the uranium metal signal. Data acquisition consisted of a 50 seconds "on-peak-zero" measurement (gas and acid blank) prior to aspiration of the sample. Upon aspiration of the sample, two half-mass unit baseline measurements were conducted (15 seconds each) within the mass range of data acquisition (232.5 to 236.5). Isotope ratios were measured for 50 scans at 10 seconds integration time. Subsequent to each analysis, washout consisted of using both 4% and 2%HNO_3 wash and rinse solutions, respectively for a combined 10–15 minutes.

The acquisition of Th isotopic measurements is essentially identical to that for uranium with the exception that data acquisition consisted of 60 scans at 5 seconds integration time, and three Faraday collectors (instead of four) were used. In addition, washout times were slightly longer at approximately 20 minutes. Data reduction was carried out using software developed by B. Ghaleb.

STABLE ISOTOPE ANALYSIS

$\delta^{18}O$ and $\delta^{13}C$ of calcite were measured on a VG Sira II mass spectrometer in an Autocarb automated carbonate analyzer at McMaster University. Isotopic ratios are reported with respect to the VPDB standard in per mil units (‰). NBS-19 is the laboratory reference standard used in this study. The precision of analyses for both isotopes is ±0.1‰.

RESULTS

U-SERIES DATES

Results of the U-series dating are presented in Table 1. Measured U concentrations are similar in all samples and high enough to permit precise estimates. Errors for the U/Th ages are reported to two standard deviations. $^{234}U/^{238}U$ ratios are also similar, suggesting that the source environment is stable. All of the Th/U ages obtained are in their correct stratigraphic order.

THE GROWTH RATES OF THE STALAGMITE

Annual growth rates were calculated by dividing the portion of calcite (mm) by the number of years that portion was deposited over (derived from Th/U dates). The calculated annual growth rates between dated sections of MF1 may provide insight into any environmental changes above the cave (Table 2 and Fig. 2). The faster the growth rate, the warmer and/or wetter the climate is above the cave (Hennig 1983; Dreybrodt 1982). We observe a distinct change in growth rate, from a maximum value near the base of c. 1.0 mm/y to a rate two orders of magnitude lower through the remainder of the history of this stalagmite. These growth rates are within the

Table 2. Annual growth rates for MF1.

Distance from Base (mm)	Time of Growth (years)	Length of Calcite (mm)	Rate of Growth (mm/yr)
0–37	36	37	1.027
38–75	246	37	0.150
76–165	2519	89	0.035
166–225	2560	59	0.023
226–270	2245	44	0.019

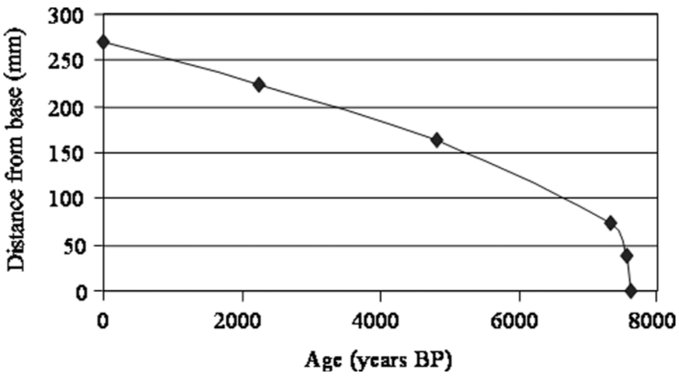

Figure 2. Variations in height with age, showing rapid deceleration of growth rates after the first 250 years.

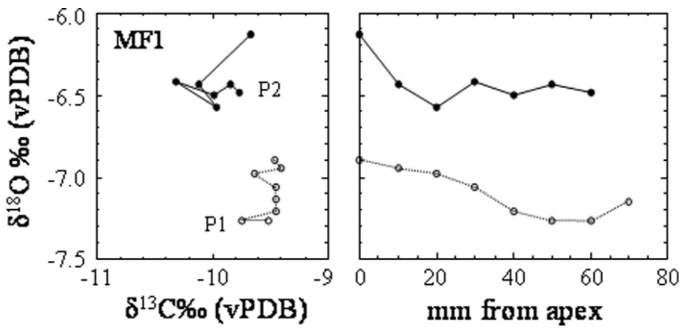

Figure 3. Hendy tests for equilibrium deposition of MF1. Equilibrium deposition is generally assumed if the $\delta^{18}O$ range is less than 0.8‰. P1 and P2 are the profiles drilled along two growth rings within MF1.

known range for speleothems from temperate climates (Baker *et al*. 1998; van Beynen 1998). Elsewhere, high growth rates have occurred during periods of temperature maxima and/or wetter conditions while growth declined during cooler/drier periods (Hennig *et al*. 1983; Railsback *et al*. 1994; Genty & Quinif 1996; Schwehr 1998; Brook *et al*. 1999; Qui *et al*. 1999). As the earlier part of the Holocene is known to have been warmer than present in the nearby Finger Lakes Region, NY (Mullins 1998), we might expect higher growth rates near the base of MF1. For the first 40 years of growth, the rates are very high, > 1 mm per year. The next 250 years also have fast growth though an order of magnitude lower. Overall, 25% of the speleothem was deposited during approximately its first 300 years, the rest over the remaining 7300 years. Such rapid growth from 7.6 to 7.3 ka for MF1 is consistent with the

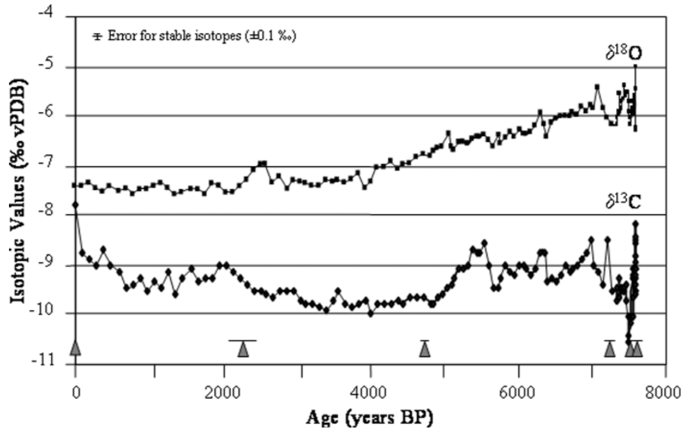

Figure 4. $\delta^{18}O$ and $\delta^{13}C$ values for MF1.

warm/wet period from 10.2 to 6.9 ky reported by Prentice *et al*. (1991) and Jackson and Whitehead (1991). Pielou (1991) places the Holocene Hypsithermal at ~7.8 ka in central New York, which is the same period that Dywer *et al*. (1996) and Mullins (1998) found high water levels at Finger Lakes of NY. These studies suggest MF1 was responding to a warm and wet climate in the NY region during this time.

Stable isotope analysis

a) *Isotopic equilibrium test*

To test whether the speleothem was deposited in isotopic equilibrium, profiles were drilled precisely along two of its growth layers. The criteria for equilibrium deposition are that there must be no correlation between the $\delta^{18}O$ and $\delta^{13}C$ for each profile or any variation in $\delta^{18}O$ greater than 0.8‰ (Hendy & Wilson 1968; Gascoyne 1992). Results shown in Figure 3 indicate that MF1 was deposited in equilibrium as both P1 and P2 in the left box obviously have no correlation between $\delta^{18}O$ and $\delta^{13}C$ (averaging at r = 0.08, significance = 0.29) and in the right hand box the profiles have $\delta^{18}O$ deviation < 0.8‰.

b) *The $\delta^{18}O$ records*

Figure 4 displays the $\delta^{18}O$ records for the MF1. Maximum $\delta^{18}O$ values are observed at the base of the stalagmite, during the period from 7.6 to 6.6 ka. $\delta^{18}O$ decreases gradually in a transition period from 6.6 to 2.2 ka and relatively stable values persist after that till present.

We know that the climate in this region around 10.2 to 6.9 ka BP was warmer than in the later Holocene. We may therefore infer that $\gamma > 0$ for this speleothem. Secondly, the trend of $\delta^{18}O$ values in MF1 during the Holocene shows that climate has been steadily cooling with only minor perturbations over the history of this deposit. The exceptionally high growth rates and $\delta^{18}O$ values at the base of the deposit suggest that growth began near the warmest phase of the Hypsithermal (Pielou 1991; Anderson *et al*. 1997).

(c) *The $\delta^{13}C$ records*

$\delta^{13}C$ values displayed in Figure 4 vary from –7.8 to

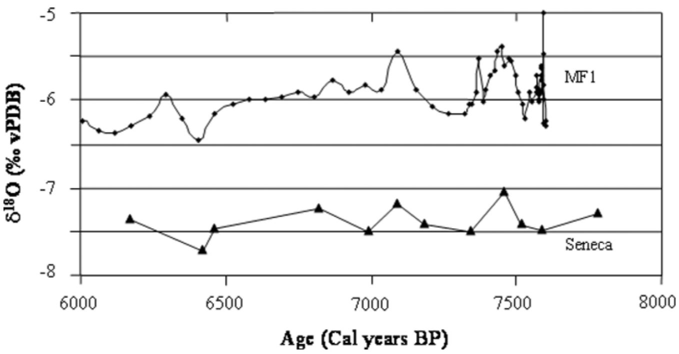

Figure 5. Comparison of MF1 δ^{18}O values with δ^{18}O values obtained from a marl record from Seneca Lake, NY (Anderson *et al.* 1997).

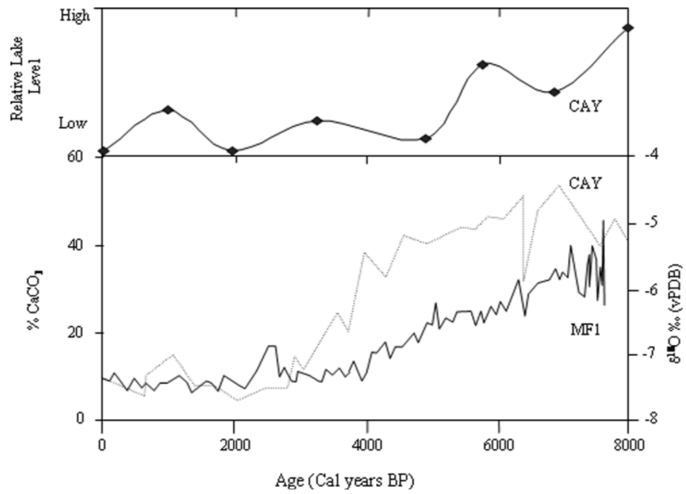

Figure 6. Comparison of MF1 δ^{18}O values with % $CaCO_3$ in sediment of Cayuga Lake and its relative lake levels (Mullins, 1998).

–10.8‰, a range of only 3 per mil. Dorale *et al.* (1998) show that for changes from C_3 to C_4 vegetation, shifts of 7‰ in δ^{13}C of speleothem calcite are required. What the McFail's δ^{13}C record shows is that forests (C_3 plants) have been the predominant vegetation above the cave with no evidence of grassland environments until very recently. Therefore, the δ^{13}C values record changes in the density of the forest above the cave. Prentice *et al.* (1991) showed that forest (C_3) existed in this region throughout the growth of this speleothem, with only tree composition changing (Jackson & Whitehead 1991). The dramatic decrease in the δ^{13}C values from 7.6 to 7.5 ka may mark the arrival of forest vegetation (*Carya*, *Ulmus*, *Fagus* and *Tsuga* species). The period from 7.0 to 5.5 with its heavier δ^{13}C values suggests a decline in the density of the forest, probably due to cooling of the regional climate. However, from 5.5 to 2.5 ka the δ^{13}C values lightened once more, even though temperatures continued to fall during most of this period. The only possible explanation for such a trend could be a wetter climate, generating more lush growth. The final 2.5 ka years show the δ^{13}C values trending towards heavier values, and as temperatures were relatively stable during this time, we suggest that climate became somewhat drier in the NY region. The steeper rise during the last 300 years may be associated with more intense deforestation by European populations (Foster & Zebryk 1993).

DISCUSSION

CLIMATE CHANGE INFERRED FROM CHANGES IN $\delta^{18}O_c$

Stalagmite MF1 appears to have begun its growth near the climatic high-point of the Holocene Hypsithermal as derived by Perliou (1991) and Anderson *et al.* (1997). This is principally suggested by the record of δ^{18}O but is also consistent with the history of the growth rate and slightly less clearly defined by the record of the δ^{13}C. It appears that $\gamma > 0$ for this speleothem: if γ remained more or less constant throughout the Holocene, then we would see a record of almost linearly decreasing temperature for the first 4,500 years of deposition (until ~3.0 ka BP) followed by stable climate until present.

The climate during the period from 7.0 ka to 7.6 ka appears to be more variable than after this phase (Fig. 4). Two distinct peaks in δ^{18}O at 7.0 ka and 7.5 ka are separated by a prominent decline suggesting a short cooling. This cooling interval and the two warm periods delineating it are also found in the Seneca Lake δ^{18}O record (Anderson *et al.* 1997), thereby showing that the climate change recorded by MF1 was regional and not highly localized (Fig. 5). The Anderson *et al.* (1997) lake record shows the same trend of falling δ^{18}O values as the MF1 record. The decreasing trend in the δ^{18}O values in both records presumably represents falling temperatures.

The variability in the record during the aforementioned period is not apparent in the rest of the record. A possible explanation can be forwarded from the growth rates. This period had the fastest growth rates and because the speleothem was sampled at regular intervals, the sampling density would be the greatest for this section. The rest of the speleothem had slower deposition rates and consequently more years were averaged when each isotopic sample was taken, thereby removing some of the variability that was evident from 7.0 to 7.6 ka. However, we would stress that even if the sampling density was the same as for later in the record, the δ^{18}O isotopic values for this period would still have shown it to be the warmest part of the Holocene as measured by MF1.

Although we have shown similarity between our δ^{18}O record for McFail's Cave and a short record from Seneca Lake for the first 1600 years of MF1's growth, we must also investigate comparisons for the other 6000 years of its growth. On a regional scale the most complete record is Mullins' (1998) record of carbonate content of sediments (marl) deposited in Cayuga Lake, NY. Figure 6 shows the comparisons with: (a) inferred lake levels, where the author states that lower levels were analogous with drier/cooler conditions, and (b) marl concentrations in the lake sediment, where Mullins attributes higher carbonate content to higher summer temperatures. Both these records show very similar trends to that of MF1, with warmer conditions in the earlier Holocene and a transition to cooler present climatic conditions by ~3.0 ka BP.

Most previous studies of the paleoclimate of the northeastern USA come from pollen records extracted from lake and bog sediments. Prentice *et al*. (1991) found that increased precipitation allowed hickory to expand in the Appalachian region 6.9 to 10.2 ka. Davis *et al*. (1980) noted that between 7.8 to 5.7 ka *Pinus*, *Strobus* and *Tsuga* were at higher elevations in the Adirondack and New England Mountains, suggesting temperatures warmer than present. MF1 does show this period as especially warm compared to the period before 7.0 ka. 5.7 to 3.5 ka saw the transition of Adirondack vegetation to more modern assemblages (Jackson & Whitehead 1991), a transition that is in accordance with the record of MF1. There was a retreat of hemlock after 3 ka that is indicative of decreasing summer temperatures (Prentice *et al*. 1991). Increasing *Picea* pollen after 2.5 ka also suggests decreasing temperatures (Jackson & Whitehead 1991). These two studies show that climate was cooler during the last 3000 years for the broader region around McFails Cave, which concurs with our $\delta^{18}O$ record.

Temperature of climatic optimum for MF1

Dorale *et al*. 1992 calculated temperatures of deposition of speleothems from their $\delta^{18}O$ values using the following equation

$$\delta^{18}O_c = (0.695T + 986.4)\exp[2780/(273.15 + T)^2]^{-0.00289} - 1000 \qquad [2]$$

where T is in °C. Using the present day annual average of 7.3 °C for Cobleskill, we obtain a present day $\delta^{18}O_c$ of 24.2‰ (SMOW) which is slightly higher than the observed value of 23.16‰. This discrepancy could be due to a difference in the T dependence of $\delta^{18}O$ (ppt) for the northeast of the USA compared with the north-central USA where Dorale *et al*. (1992) conducted their study. Rozanski *et al*. (1993) suggest a value of 0.58‰/°C rather than 0.695. Changing this value results in agreement between the calculated and observed present day $\delta^{18}O_c$ values. Using this revised equation it is possible to determine the temperature at McFail's Cave at the climatic optimum. Our average $\delta^{18}O_c$ value for c. 7.6 ka is 24.89‰ SMOW. Using the modified version of Equation [2] we obtain a temperature of 12.5°C, which would be ~5°C warmer than present day temperatures. We can compare this result with that of Prentice *et al*. (1991) who provide snapshots at 3 ka increments using isotherms inferred from pollen data for the northeastern USA to reconstruct January and July temperatures. Plotting our site on their maps for 6.9 ka BP, which is their warmest snapshot during the Holocene, we obtain a mean annual temperature of 11.8°C, which is 4.3°C warmer than present, in good agreement with our estimate.

Another factor that may have influenced the $\delta^{18}O$ of calcite through the Holocene is a change in the seasonal distribution of precipitation. Using weather records for nearby Cobleskill, NY, we see that at present, precipitation is distributed approximately uniformly throughout the year. We also note that at present, the maximum range in temperature between winter and summer is about 28°C; assuming an isotopic temperature dependence of 0.58‰/°C, this represents a maximum W-S difference in $\delta^{18}O$ (ppt) of c. 16‰. More realistically, the difference in average winter and summer temperatures (October-March vs. April-September) is 9°C, which corresponds to a difference in $\delta^{18}O$ (ppt) of 5.2‰. For each 1% increase in the relative proportion of summer rain relative to winter snow + rain, we would increase $\delta^{18}O$ of average precipitation by 0.05‰. It is conceivable that there could have been as much as 10% more summer precipitation at the Hypsithermal due to retreat of the polar front (Dywer *et al*. 1996) and decrease in winter storm activity. This could have accounted for an increase of about 0.5‰ in average annual precipitation which would be reflected in $\delta^{18}O$ of speleothem.

While the seasonal distribution of rainfall may have affected the $\delta^{18}O$ values of the speleothem, the source of the rainfall may also have changed. Evidence for such a change is provided by the interpretation of eolian deposits in the Central United States (Forman *et al*. 1995). As an explanation for arid conditions in their records in the mid-Holocene, the authors suggested that the Jet Stream had moved to higher latitudes, shifting the location of the Bermuda High which then funneled moisture to the Northeast and not the Mid-West. Around 3.0 ka BP the Jet Stream retreated southward, and consequently the Bermuda High channeled moisture from the Gulf of Mexico into the interior of the continent. McFail's Cave is located in this transitional zone of the Jet Stream, and therefore its movement may have changed the source of the rainfall that contributed to the seepage of the drip waters feeding MF1. However, this movement would also have affected the temperature of the air above and consequently within the cave as well. A Jet Stream to the north of the cave would have brought warmer temperatures and rainfall from a source area with heavier $\delta^{18}O$ values to McFail's Cave, while its retreat south would have had the opposite effect. The shifts in the location of the Jet Stream as identified by Forman *et al*. (1995) are recorded in our $\delta^{18}O$ data. However, the Forman article's aim was not to provide precise locations of the Jet Stream, and its exact position is not known throughout the Holocene. This makes quantification of the effect these shifts may have on our isotopic record nearly impossible. All that can be said with any confidence is that our estimate of temperature at 7.6 ka BP can be considered a maximum value.

Climate change inferred from growth rate

Maximum growth rates for MF1 were from 7.3 to 7.6 ka. This is consistent with our $\delta^{18}O$ record, being indicative of warmer regional temperatures that would lead to faster growth rates (Dreybrodt 1982). However, the growth rates at the inception of deposition of this stalagmite are exceptionally high and suggest the occurrence of extraordinary climatic conditions at this time, possibly a combination of high temperatures and high rainfall. A warm climate is shown in the higher lake levels at Cayuga Lake, NY (Mullins 1998) providing evidence for higher rainfall in this region during this period.

Another possibility could be the presence of denser forest cover, with greater respiration increasing the CO_2 dissolved in the percolation waters. Denser forest cover would be expected to lead to lower $\delta^{13}C$ values; such a trend is observed briefly around 7.5 ka BP.

CONCLUSIONS

The climatic interpretation derived from records of both $\delta^{18}O$ and speleothem deposition rate at McFail's Cave shows a warmer and probably wetter period than at present from 7.0 to 7.6 ka. From 7.0 to 3.0 ka we see a steady progression towards the present cooler temperatures. The growth rates did not change markedly over this period, suggesting little change in precipitation or vegetational cover above the cave. From 2 to 0 ka there is little change in temperatures for this region. These findings regarding temperature are in general agreement with other paleoclimatic studies for this region. There is a brief warm spell centered on 2.5 ka that punctuates the cooler present temperatures. The highest $\delta^{18}O$ values were observed at the start of deposition of MF1, indicating the warmest temperatures, ~5°C warmer than present, though a shifting Jet Stream bringing isotopically heavier precipitation to the New York region would moderate this value somewhat. MF1 may however not record the time of peak temperatures of the Hypsithermal; a longer stable isotope profile of lake sediments from the region extending to 8.9 ka by (Anderson *et al.* 1997) also shows declining $\delta^{18}O$ (and therefore temperature) values, indicating that the maximum of the Hypsithermal may precede our inferred climatic optimum from MF1.

The $\delta^{13}C$ record showed some variability, with long-term changes in the density of the forest vegetation above the cave over this period. A spike of lower $\delta^{13}C$ values at ~7.5 ka shows an increase in productivity of the forest above the cave with the arrival of certain tree species (Prentice *et al.* 1991) which corresponds to the period of enhanced growth rate of stalagmite MF1. There was also a longer but less pronounced increase in forest density from 5.5 to 2.5 ka BP, suggesting increases in precipitation rates for the region.

ACKNOWLEDGEMENTS

This research has been made possible by NSERC grants to Schwarcz and Ford. Special thanks to Nicki Robinson for processing the U-series dates and Martin Knyf for assistance in the stable isotope analysis. We would also like to thank H.T. Mullins for his comments that helped improve this manuscript.

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